AFOSR - TR - 72 - 0373

AD737135

FINAL REPORT

31 December 1971

RESEARCH IN SEISMOLOGY

PREPARED FOR

Geophysics Division

Air Force Office of Scientific Research

Arlington, Virginia 22209

SEE AD706872

by

Department of Earth and Planetary Sciences

Massachusetts Institute of Technology

Cambridge, Massachusetts 02139

SEE AD706872

Sponsored by

Advanced Research Project Agency

Nuclear Test Detection Office

Project Vela-Uniform

ARPA Order 292-75

65

DOCUMENT CONTROL DATA - R & D

(Security classification of title, body of abstract and indexing annotation must be entered when the overall report is classified)

1. ORIGINATING ACTIVITY (Corporate author) Massachusetts Institute of Technology Department of Earth and Planetary Sciences Cambridge, Massachusetts 02139	2a. REPORT SECURITY CLASSIFICATION UNCLASSIFIED 2b. GROUP
3. REPORT TITLE Research in Seismology	
4. DESCRIPTIVE NOTES (Type of report and inclusive dates) Scientific:.......Final	
5. AUTHOR(S) (First name, middle initial, last name) Frank Press M. Nafi Toksöz Keiiti Aki	
6. REPORT DATE 31 December 1971	7a. TOTAL NO. OF PAGES 65 / 7b. NO. OF REFS 50
8a. CONTRACT OR GRANT NO. AF 49(638)1632 b. PROJECT NO. AO-292-75 c. 62701D d.	9a. ORIGINATOR'S REPORT NUMBER(S) 9b. OTHER REPORT NO(S) (Any other numbers that may be assigned this report) AFOSR - TR - 72 - 0873
10. DISTRIBUTION STATEMENT Approved for public release; distribution unlimited.	
11. SUPPLEMENTARY NOTES TECH, OTHER	12. SPONSORING MILITARY ACTIVITY Air Force Office of Scientific Research (NPG) 1400 Wilson Boulevard Arlington, Virginia 22209

13. ABSTRACT

A brief summary of the research efforts under this contract is provided. The areas of investigation fall under five categories: 1) Seismic wave propagation in heterogeneous media, where finite difference and other numerical techniques are applied to non-uniform geometries, 2) Seismic source theory, and the determination of source parameters and focal depth from spectra of Rayleigh and Love waves, 3) Study of structure, properties and heterogeneities of the earth's interior, from seismic body and surface waves and generation of average earth models, 4) Utilization of seismic arrays, and automatic schemes of phase identification and event verification from the LASA detection logs, and 5) Development and performance of long-period tiltmeters. A comprehensive list of articles published on the above subjects is provided.

DD FORM 1 NOV 65 1473

UNCLASSIFIED

Security Classification

AVAILABILITY

Qualified requestors may obtain additional copies from the Defense Documentation Center, all others should apply to the Clearinghouse for Federal Scientific and Technical Information.

Distribution of this document is unlimited.

Department of Earth and Planetary Sciences
Massachusetts Institute of Technology
Cambridge, Massachusetts 02139

RESEARCH IN SEISMOLOGY

Final Report

to

Air Force Office of Scientific Research

1 November 1965 - 31 March 1971

ARPA Order No. 292 Amendment 75

Project Code No. AO-292-75

Name of Contractor - Massachusetts Institute of Technology

Date of Contract - 1 November 1965

Amount of Contract - $917,398.00

Contract No. AF 49(638)-1632

Contract Termination Date - 31 March 1971

Project Scientists - Frank Press; 617/864-6900, ext. 3382
M. Nafi Toksöz; 617/864-6900, ext. 6382
K. Aki; 617/864-6900, ext. 6397

Short Title of Work - Research in Seismology

ABSTRACT

A brief summary of the research efforts under this contract is provided. The areas of investigation fall under five categories: 1) Seismic wave propagation in heterogeneous media where finite difference and other numerical techniques are applied to non-uniform geometries, 2) Seismic source theory, and the determination of source parameters and focal depth from spectra of Rayleigh and Love waves, 3) Study of structure, properties and heterogeneities of the earth's interior from seismic body and surface waves and generation of average earth models, 4) Utilization of seismic arrays, and automatic schemes of phase identification and event verification from the LASA detection logs, 5) Development and performance of long period tiltmeters. A comprehensive list of articles published on the above subjects is provided.

TABLE OF CONTENTS

I. INTRODUCTION

In this final report we will briefly summarize the work done under the contract during the past five years. As the title of the contract indicates, this project was intended as a study of broad aspects of seismology and the understanding of the specific problems within the general development. The studies cover the elastic and non-elastic properties of the earth and the nature of earthquake and explosive sources. Greatest efforts were made to go beyond the classical techniques in studying general problems.

The areas of investigations can be combined under five general headings:

1. Seismic wave propagation studies
2. Source mechanism and focal depth
3. Earth structure and properties
4. Seismic array studies
5. Instrument developments

Because of length limitations, we cannot discuss in detail all the work done under each heading. Instead we introduce a brief summary and list the published papers which describe the completed studies. The unpublished results are discussed in greater detail.

At the end of the report a list of published papers and the titles of graduate theses supported under this contract is given.

II. WAVE PROPAGATION STUDIES

In this area the efforts have been concentrated in studying seismic wave propagation in heterogeneous media. In the past much has been learned about the earth's interior and the seismic source from the interpretation of data in terms of laterally homogeneous models where properties depended only on depth. Increased data on the lateral variations of properties and structure, not only in the crust but also in the mantle, have made it necessary to consider seismic surface and body wave propagation in heterogeneous media.

By the nature of the problem, such studies can best be carried out by numerical methods. Finite-difference techniques, perturbation methods, and statistical techniques were utilized in these investigations. Problems of interest included surface wave propagation in media with non-parallel boundaries, body waves incident on non-plane boundaries, and scattering from randomly distributed heterogeneities.

For the propagation of body and surface waves in media with non-plane boundaries (such as sedimentary basins and the crustal structure under the Large Aperture Seismic Array (LASA) in Montana), finite-difference techniques were used. For SH waves and Love waves the effects of heterogeneities were found to be significant. Although the results could have been predicted intuitively for smooth structures, for

models with sharp boundaries or heterogeneities of the order of a wave length the effects become very complicated because of scattering, wave conversion, and interference phenomena.

For seismic sources in regions of downgoing slabs, both ray theory and wave theory formulations have been developed. Very strong asymmetry in radiated seismic body waves are observed. Strong travel-time anomalies, amplitude variations with azimuth and distance from earthquakes and explosions such as Longshot and Milrow can be explained by the lateral variation of velocities in and around the downgoing slab under island arcs.

The splitting of the eigenfrequencies of torsional free oscillations of the earth due to lateral heterogeneities is studied by perturbation theory. The perturbations are expanded in spherical harmonics $[\delta\rho_s^t Y_s^t(\theta,\phi),\ \delta\mu_s^t Y_s^t(\theta,\phi)]$ and it is found that odd terms (s=odd) produce no splitting to first order. For example, there would be no splitting if the earth had a continental hemisphere and an oceanic hemisphere. Also, heterogeneities such that $s>2\ell$ will not affect the eigenfrequencies of modes of order ℓ. For large ℓ this means that free oscillations will not be affected by heterogeneities with scale lengths less than half the wavelength of the free oscillation. The amount of splitting is computed for earth models. The interpretation of high angular-order modes as long-period surface waves propagating

along great-circle paths are analyzed and some limits are established for the technique of phase and group velocity averaging along great-circle paths.

Publications

Aki, K., Analysis of the seismic coda of local earthquakes as scattered waves, J. Geophys. Res., 74, 615-631, 1969.

Aki, K. and K. Larner, Surface motion of a layered medium having an irregular interface due to incident plane SH waves, J. Geophys. Res., 75, 933-954, 1970.

Aki, K. and Y.B. Tsai, The mechanism of Love wave excitation by explosive sources, (submitted to J. Geophys. Res., 1971).

Boore, D., Finite Difference solutions to the equation of elastic wave propagation, with application to Love waves over dipping interfaces, Ph.D. thesis, M.I.T., 1969.

Boore, D., Effect of higher mode contamination on measured Love wave phase velocities, J. Geophys. Res., 74, 6612- , 1969.

Boore, D., Love waves in nonuniform wave guides: Finite difference calculations, J. Geophys. Res., 75, 1512- , 1970.

Boore, D., K. Larner and K. Aki, Comparison of two independent methods for the solution of wave-scattering problems: response of a sedimentary basin to vertically incident SH waves, J. Geophys. Res., 76, 558-569, 1971.

Frasier, C., Discrete time solution of plane P-SV waves in a plane layered medium, Ph.D. thesis, M.I.T., 1969.

Frasier, C., Discrete time solution of plane P-SV waves in a plane layered medium, Geophysics, 35, 197-219, 1970.

Larner, K., Near-receiver scattering of teleseismic body waves in layered crust-mantle models having irregular interfaces, Ph.D. thesis, M.I.T., 1969.

Madariaga, R., Study of free oscillations of the laterally heterogeneous Earth models, abstract, program for Cordilleran Section, 67th Annual Meeting at Riverside, Calif.,Geol. Soc. Am., 1971.

Madariaga, R., Free oscillations of the laterally heterogeneous earth, Ph.D. thesis, M.I.T., 1971.

Toksöz, M.N., J.W. Minear and B.R. Julian, Temperature field and geophysical effects of a downgoing slab, J. Geophys. Res., 76, 1113-1138, 1971.

III. SOURCE MECHANISM AND FOCAL DEPTH STUDIES

In the study of the earthquake source mechanism and focal depth, both body and surface waves, with special emphasis on the latter, have been used. A dislocation model of an earthquake source was used for determination of fault plane geometry, seismic moment, fault length, and focal depth. The method is based on the theory of normal mode excitation by a seismic source in a vertically heterogeneous medium. The spectra of seismic waves recorded at distant stations are corrected for propagation effects, computed on the basis of structural models. Then the corrected spectra are interpreted for source parameters.

From the spectra of Rayleigh and Love waves at a number of stations, in addition to the dislocation parameters and fault depth, the source time function can be obtained. The effects of these parameters on the surface and body wave magnitudes (M_S and m_b), and hence on the seismic discrimination problem, are quite significant. Since the source dimensions and source time function control the spectral shapes, they also determine the relative amplitudes of body and surface waves measured at 1 Hz and 0.05 Hz, respectively. In addition to above, source depth is an important parameter in the spectral shape of surface waves, as well as being a discriminant in itself. Thus a method of accurate determination of focal depths of shallow events has been

sought. The surface waves have proved to be most sensitive to focal depth, if the source mechanism and source time function is known, or kept constant.

Utilizing Rayleigh and Love waves, source characteristics of both underground nuclear explosions and many earthquakes are investigated. These are discussed in great detail in the publications listed below. One aspect of the underground explosion source function which complicates the above procedure is the release of tectonic strain energy by the explosions. Even though strong Love waves are generated by some explosions, the M_S - m_b discriminant is not seriously effected.

Dislocation models were applied to an ensemble of earthquakes by invoking a "similarity" of large and small earthquakes. A scaling law of seismic spectra, derived on the basis of an "ω-squared" model, have been used in the discussion of various problems including the theoretical basis for the M_S - m_b discriminant.

Publications

Aki, K., Scaling law of seismic spectrum, J. Geophys. Res., 72, 1217-1231, 1967.

Aki, K., Seismological evidences for the existence of soft thin layers in the upper mantle under Japan, J. Geophys. Res., 73, 585-594, 1968a.

Aki, K., Seismic displacements near a fault, J. Geophys. Res., 73, 5359-5376, 1968b.

Kanamori, H., The Alaska earthquake of 1964: Radiation of long-period surface waves and source mechanism, J. Geophys. Res., 75, 5029-5040, 1970.

Kanamori, H., Synthesis of long-period surface waves and its application to earthquake source studies - Kurile Islands Earthquake of October 13, 1963, J. Geophys. Res., 75, 5011-5027, 1970.

Toksöz, M.N., Radiation of seismic waves from underground explosions, Proceedings of the Vesiac conference on the current status and future prognosis of shallow seismic events, VESIAC Report, Willow Run Laboratories, The University of Michigan, 65-84, 1967.

Toksöz, M.N., Crustal effects on long-period chirp filters, Proceedings of the Woods Hole Conference on Seismic Discrimination, Vol. I, 20-23 July 1970.

Toksöz, M.N. and H.H. Kehrer, Underground nuclear explosions:

tectonic utility and dangers, Science, 173, 230-233, 16 July 1971.

Tsai, Y.B., Determination of focal depths of earthquakes in the mid-oceanic ridges from amplitude spectra of surface waves, Ph.D. thesis, M.I.T., 1970.

Tsai, Y.B. and K. Aki, Simultaneous determination of the seismic moment and attenuation of seismic surface waves, Bull. Seism. Soc. Am., 59, 275-287, 1969.

Tsai, Y.B. and K. Aki, Precise focal depth determination from amplitude spectra of surface waves, J. Geophys. Res., 75, 5729-5743, 1970.

Tsai, Y.B. and K. Aki, Amplitude spectra of surface waves from small earthquakes and underground nuclear explosions, J. Geophys. Res., 76, 3940-3952, 1971.

IV. EARTH STRUCTURE AND PROPERTIES

The elastic and anelastic properties of the earth are both radially and laterally heterogeneous. A great deal of effort has gone into examining the detailed variations of these properties, with particular emphasis on understanding the causes of the variations and their relationship to other geophysical measurements and to tectonic patterns. Discussion of this work is divided below into three sections: (1) gross earth structure, including seismic velocity and density, (2) lateral heterogeneities, including velocity, density, and Q, and (3) the important effects of these findings, particularly those on Q, on seismic wave magnitudes and the discrimination problem.

1) Earth structure

Various techniques have been used to infer the variation with depth of seismic velocity and density in the earth. With a large array, the slope $dt/d\Delta$ of the travel-time curve can be measured directly. Using LASA, observations of $dt/d\Delta$ have been inverted to obtain P- and S-wave velocities in the lower mantle (Chinnery and Toksöz, 1967; Toksöz et al., 1967; Fairborn, 1968,1969) and estimates of P-wave velocity in the upper mantle (Wolfe, 1969). An important finding was the discovery of several sharp increases in velocity with depth in the lower mantle, suggesting that the lower mantle is not homogeneous.

With large, efficient computers, the generation of velocity-density models for the earth that satisfy the entire range of seismic data, as well as mass and moment of inertia, is a simple process. The problem is inherently non-unique, however. A successful attack on the question of non-uniqueness is the Monte Carlo method, in which randomly generated earth models are tested against observations. The Monte Carlo technique, which offers freedom from bias and which emphasizes the traits common to successful models, has been extensively applied to the problem of earth structure. Highlights of the conclusions include constraints on the composition of mantle and core and insights into the nature and thickness of lithosphere and asthenosphere (Press, 1968,1969,1970a,b; Press and Kanamori, 1970).

2) Lateral variations

Both the elastic and anelastic properties of the crust and upper mantle vary with tectonic setting. The phase and group velocities of long-period surface waves are different for oceans, shields, and tectonically active regions. Inversion of the dispersion data requires that significant differences in shear velocity among the various regions persist to depths of at least 400 km (Toksöz et al., 1967).

Even more regionally variable is the seismic quality factor Q, particularly within the low-Q, low-velocity

asthenosphere. These lateral variations may be observed both in the differential attenuation of teleseismic P-and S-waves (Solomon and Toksöz, 1970) and in the attenuation of surface waves of period sufficiently long so that a sizeable fraction of the wave energy propagates in the asthenosphere (Tsai and Aki, 1969; Solomon, 1971a).

Because the asthenosphere, at least in tectonically active areas, is almost certainly partially melted, theoretical considerations suggest that Q at that depth should be a function of frequency. This is, in fact, required by the data for western North America: attenuation and velocity measurements of various sorts, in all spanning three decades in frequency, can be fit only by postulating a Q that depends both on depth and frequency (Solomon, 1971a,b). The precise dependence of Q on frequency is governed by properties of the partial melt such as melt viscosity and volume concentration of melt. All of the lateral variation of Q in the asthenosphere may be explained by relatively modest lateral changes in temperature (Solomon, 1971b).

Among the most dramatic lateral variations in such quantities as velocity and Q occur at island arcs, where the relatively cool lithosphere is being underthrust into the mantle. Seismic velocities calculated from realistic temperature models for downgoing lithospheric slabs (Minear and Toksöz, 1970a,b; Toksöz et al., 1971) predict that the

slab will strongly affect the travel-time and propagation direction of body waves from nearby sources. This has important implications for determination of location and source mechanism for seismic events in such regions (Toksöz et al., 1971).

Spherical harmonic expansion and correlation of such globally distributed geophysical data as seismic travel-time anomalies, heat flow, surface topography and gravity demonstrate both the interrelationships of these various quantities and the nature of the global variations (Toksöz and Arkani-Hamed, 1967; Arkani-Hamed and Toksöz, 1968; Toksöz et al., 1969; Arkani-Hamed, 1969,1970). A major conclusion of such studies is that lateral variation of such properties as density extends well into the lower mantle.

3) Some effects on magnitudes and discrimination

The nature of heterogeneities in the earth must be well understood before seismic wave amplitudes may be corrected for propagation effects to estimate source parameters. Foremost among the relevant earth properties is Q, which controls both the absolute amplitudes and spectral shape of seismic waves.

Lateral variation of Q can appreciably shift the $M_S - m_b$ relationship. In particular, in tectonically active areas such as parts of western North America, m_b is generally lowered by several tenths of a magnitude unit, relative to

events of similar M_S in more stable regions, due to abnormally high attenuation in the upper mantle of the tectonic region (Solomon et al., 1970; Ward and Toksöz, 1971). The reason is simple. Because surface waves in the period range used to define M_S propagate principally in the high-Q lithosphere, and because scattering at ocean-continent boundaries is generally small at these periods, determination of M_S is relatively insensitive to path. For body waves that have propagated through the low-Q asthenosphere in which Q is strongly variable from one region to another, m_b is strongly path dependent.

In any amplitude measurements that cover a broad frequency band, some account needs to be made of the regional variations and frequency dependence of Q in order to extract the maximum possible information about the source. In particular, determination of source depth from surface wave spectral shape (see section III) and of source dimensions and/or time function from body wave spectra must be based on an accurate knowledge of Q in the earth.

Publications

Aki, K., Seismic displacements near a fault, J. Geophys. Res., 73, 585-594, 1968a.

Arkani-Hamed, J., Lateral variations of density in the earth's mantle, Ph. D. Thesis, M.I.T., 1969.

Arkani-Hamed, J., Lateral variations of density in the mantle, Geophys. J. R. astr. Soc., 20, 431-455, 1970.

Arkani-Hamed, J. and M.N. Toksöz, Analysis and correlation of geophysical data, Supplemento al Nuovo Cimento, Ser. I, 6, 22-66, 1968.

Boore, D.M. and M.N. Toksöz, Rayleigh wave particle motion and crustal structure, Bull. Seism. Soc. Am., 59, 331-346, 1969.

Chinnery, M.A. and M.N. Toksöz, P-wave velocities in the mantle below 700 km, Bull. Seism. Soc. Am., 57, 199-226, 1967.

Fairborn, J. W., Mantle P and S wave velocity distributions from dT/dΔ measurements, Ph. D. Thesis, M.I.T., 1968.

Fairborn, J.W., Shear wave velocities in the lower mantle, Bull. Seism. Soc. Am., 59, 1983-1999, 1969.

Kanamori, H. and F. Press, How thick is the lithosphere? Nature, 226, 330-331, 1970.

Minear, J.W. and M.N. Toksöz, Thermal regime of a downgoing slab and new global tectonics, J. Geophys. Res., 75, 1397-1419, 1970a.

Minear, J.W. and M.N. Toksöz, Thermal regime of a downgoing slab, Tectonophysics, 10, 367-390, 1970b.

Press, F., Earth models obtained by Monte Carlo inversion, J. Geophys. Res., 73, 5223-5234.

Press, F., The sub-oceanic mantle, Science, 165, 174-176, 1969.

Press, F., Earth Models Consistent with Geophysical Data, Phys. Earth Planetary Interiors, 3, 3-22, 1970.

Solomon, S.C., Seismic wave attenuation and the state of the upper mantle, Ph. D. Thesis, M.I.T., 1971a.

Solomon, S.C., Seismic-wave attenuation and partial melting in the upper mantle of North America, J. Geophys. Res., in press, 1971b.

Solomon, S.C. and M.N. Toksöz, Lateral variation of attenuation of P and S waves beneath the United States, Bull. Seism. Soc. Am., 60, 819-838, 1970.

Solomon, S.C., R.W. Ward, and M.N. Toksöz, Earthquake and explosion magniutdes: the effect of lateral variation of seismic attenuation, Proceedings of the Woods Hole Conference on Seismic Discimination, Vol. I, 20-23 July 1970.

Toksöz, M.N. and J. Arkani-Hamed, Seismic delay times: correlation with other data, Science, 158, 783-785, 1967.

Toksöz, M.N., J. Arkani-Hamed, and C.A. Knight, Geophysical data and long wave heterogeneities of the earth's mantle, J. Geophys. Res., 74, 3751-3770, 1969.

Toksöz, M.N., M.A. Chinnery, and D.L. Anderson, Inhomogeneities in the earth's mantle, Geophys. J. R. astr. Soc., 13, 31-59, 1967.

Toksöz, M.N., J.W. Minear, and B.R. Julian, Temperature field and geophysical effects of a downgoing slab, J. Geophys. Res., 76, 1113-1138, 1971.

Tsai, Y.B. and K. Aki, Simultaneous determination of the seismic moment and attenuation of seismic surface waves, Bull. Seism. Soc. Am., 59, 275-287, 1969.

Ward, R.W. and M.N. Toksöz, Causes of regional variation of magnitude, Bull. Seism. Soc. Am., 61, 649-670, 1971.

Wolfe, J.C., dt/dΔ measurement of short period P waves in the upper mantle for the western North America using LASA, S.M. Thesis, M.I.T., 1970.

V. SEISMIC ARRAY STUDIES

With the installation of the Large Aperture Seismic Array (LASA), a new capability was added to seismological studies. Availability of seismic data in digital form from a series of sensors has opened new areas of study. Under this contract we have been involved in three separate investigations: (1) Measurement of apparent slowness of body waves ($dt/d\Delta$) and the interpretation of these in terms of detailed earth structure, (2) Frequency-wave number study of the nature of seismic noise and microseisms, and (3) Evaluation of the automatic event detection capability of arrays.

1) $dt/d\Delta$ measurements

The $dt/d\Delta$ measurements for seismic P and S waves have greatly enhanced our ability to determine the finer features of the earth's mantle. These results were discussed in greater detail in the previous section.

2) Frequency-wave number spectra and nature of microseisms

Utilizing data from LASA, microseisms have been studied by computing the frequency-wave number spectra. Wave types, modes of propagation and source regions have been identified. At LASA short-period (0.5 - 2.0 sec) microseisms consist primarily of coherent body waves, generated at the oceanic storm centers. At intermediate periods (2 - 5 sec) higher-mode surface waves dominate. At still longer periods, the

microseisms are made of fundamental mode Rayleigh waves and some Love waves.

3) <u>Automatic event detection</u>

Since April 1969 detection of seismic signals has been done automatically and on-line by the deployment of 300 pre-steered beams. Each time the signal to noise ratio goes above the 10 db threshold in any beam the event detector records a detection and groups it together with any other beam detections occurring at the same time. The output of the detector is a list of detection groupings with their times, intensity, and beam number of maximum intensity. There are about 160 detection groups a day of which only 25 are genuine signals. This study was undertaken to find how one can separate the signals from the false alarms and how one can classify the signals into first arrivals, P or PKP and the later phases PcP, ScP, etc. using some automatic scheme.

If processing of earthquake data could be done automatically, then earthquake reporting would become more uniform and standardized with time. Furthermore, more phases would be picked up since the weak phases would not be ignored if they happened to fit travel-time tables and come from the same azimuth as the earthquake. If seismic data is available only from one localized array then information from later phases is very useful in improving estimates of the

earthquake location. In some cases the detection of a later phase would confirm the interpretation of a first arrival. Later phases may contain relevant information about the focal mechanism.

In the first figure the percent detection by LASA of Coast and Geodetic Survey (CGS) reported events of magnitude m_b and greater is plotted for two distance ranges. The earthquakes missed when the LASA system was not operating were not taken into account. The number of events missed in the distance range 30 to 90 degrees appears to be independent of magnitude. Since from experience it is known to be highly unlikely for LASA to miss an event of magnitude 4.5 or larger in this distance range, it can be concluded that LASA has the capability to detect all events that CGS lists in the 30 to 90 degrees distance range. Furthermore, there are many more events that LASA reports but are unlisted in the CGS bulletins. At larger distances near and beyond the shadow zone, LASA misses not more than 40% of the events reported by CGS. A magnitude dependence can be seen. Thus it follows that though a single array does not excel in location capabilities it certainly has exceptionally good detection capabilities.

In developing the automatic phase identifier the guiding principle was to use the same ideas and methods that an analyst would employ. Given any pair of detections the

parameters of the detections, namely velocity, azimuth, signal intensity, and time interval between detections must satisfy a certain relationship (with some statistical error) in order to be classified as a certain pair of phases related to the same earthquake.

Phase Identifier - The input to the phase identifier is a list of detection times in chronological order with their signal strength and beam number of maximum intensity. The output is the interpretation of these detections, P, PKP, noise or later phase. The phase identifier will recognize the following later phases: PcP, ScP, PP, SKS, PKKP, and P'P'. In addition it will find the first arrival associated with the later phase. The phase identifier cannot pick out a later phase in the detection log if the first arrival associated with the later phase was missed. Thus it tends to miss some later phases coming from events around 110 degrees because the first arrival was undetected.

Due to the fact that the input information is inexact, the statistical approach was used in classifying a signal. The design criterion was to minimize the misclassification error hence the optimum scheme is the maximum likelihood test. Using a set of detections in which the classification of the detections was known from outside sources such as the CGS catalog and Seismic Array Analysis Bulletins (SAAC), an estimate was made of the likelihood function given a set of

observations and a specific interpretation.

$$\Lambda_i = Pr(NBM1, MSTA1, \Delta T, NBM2, MSTA2 | H_i) \qquad (1)$$

Λ_i is the probability of observing the first arrival in beam NBM1 with intensity MSTA1 followed by a later arrival ΔT seconds later in NBM2 and with intensity MSTA2 assuming that the detection pair are identified by hypothesis H_i (e.g. PKP-PKKP). The likelihood function was approximated and stored by the phase identifier so that given any detection pair with their parameters and given an interpretation, the likelihood of observing these parameters for that interpretation could be estimated. The likelihood function is computed for all possible detection pairs separated by less than 30 minutes and for all possible interpretations, and then the interpretation with the largest likelihood is chosen. The bulk of this section will deal with the problem of how to estimate Λ_i.

It is not practical to work with the parameters NBM1, MSTA1, ΔT, NBM2, MSTA2 since they are generally not independent of each other. It is very desirable if Λ_i could be approximated by a set of separable parameters, i.e.

$$\Lambda_i = Pr(S1|H_i)Pr(S2|H_i)Pr(S3|H_i) \qquad (2)$$

where $Pr(S1|H_i)$ is the probability of observing parameter S1 given the hypothesis H_i. For the set of new parameters the following were adopted:

$$S1 = DIS(NBM1) - DIS(NBM2|H_i)$$

$$S2 = DIS(NBM1) - DIS(\Delta T|H_i)$$

$$S3 = AZ(NBM1) - AZ(NBM2|H_i)$$

$$MSTA1 = MSTA1$$

$$\Delta MSTA = MSTA1 - MSTA2$$

where $DIS(NBM2|H_i)$ is the distance associated with NBM2 given it is some phase H_i, and AZ is the azimuth associated with the respective beam and hypothesis. $DIS(\Delta T|H_i)$ is the distance inferred from travel-time tables. (Depth corrections were unnecessary; see Figures 2 and 3.) These parameters should be approximately independent of each other.

These parameters have another important property. If the specific hypothesis H_i is correct then the two detections are associated with the same event and S1, S2, and S3 should cluster around zero. On the other hand if the interpretation is wrong then S1, S2, and S3 can take any value.

In actual cases S1, S2 and S3 can have a considerable departure from zero under a correct interpretation. Figure 4 plots CGS epicenters of specific phases coming in specific beams. The scatter can be attributed to incc erencies among the waves from the different subarray and the fact that with a finite number of beams it is unlikely for the phase

to fall directly in a beam. The distances and azimuths associated with the transformation were determined using a training set of detections, a set of 2000 detections which were identified from CGS and SAAC bulletins. There were many beams and phases for which no detections were found in the training set. For instance there were only 30 beams in which P'P' detections were observed. If for a specific phase no detection in the training set was observed in the beam then that phase was categorically rejected in the present implementation of the phase identifier. Thus the phase identifier would tend not to classify phases coming from aseismic regions. (This restriction could be removed for strong signals.)

In the upper half of Figure 5 the marginal distribution of S1 is shown for correctly identified phases shown in the lower half. (A synthetic detection log described in the next section was used to get the lower distributions.) The lower distributions are not uniform since there was a certain amount of screening before considering an interpretation. For example the phase identifier would not consider the P-P'P' interpretation if ΔT was less than 1400 seconds.

These distributions summarize the usefulness of the parameters S1, S2, and S3. S1 and S2 are necessary to distinguish the later phases since they are the only parameters which depend strongly on the interpretation (S3 only separates

PcP, ScP, SKP, PP interpretations from PKKP and P'P').
S3 shows the best contrast under the two hypotheses and is very useful in deciding that the two signals came from the same earthquake.

The covariance matrix for S1, S2 and S3 is shown in Table 1 for correct and incorrect identifications. The cross terms were small enough to neglect.

The remaining parameters, MSTA1 and MSTA are principally used to distinguish a genuine signal from a false alarm. The MSTA is the maximum short term average, averaged over 1.8 seconds of filtered, rectified clipped beam data sampled 10 times a second. It is measured in so called quantum units. In Figure 6 MSTA is plotted versus magnitude of shallow events at a distance between 30 and 60 degrees. (The attenuation factor is fairly constant in this range.) There appears to be a better correlation suggesting that MSTA is probably more logarithmically related to the signal amplitude; however, the scatter is so large that one should not attempt to estimate the magnitude from MSTA.

In Figure 7 the cumulative frequency of detections versus MSTA is plotted for all detections in the time interval April 13, 1969 to October 1, 1969. In the range MSTA between 400 and 1000 quantum units, the log frequency distribution is linear. Above 1000 the signals in the different subarrays are clipped. For small MSTA the number

of detections increases abnormally. It was observed that the strong signals seem to be confined to a limited set of beams corresponding to seismic regions while the weak signals occur uniformly over both seismic and aseismic regions. In Figure 8 cumulative distribution of detections is plotted versus MSTA for two sets of 36 beams, the first set has 61 percent of the detections with MSTA > 250 while the second set has less than one percent of the detections with MSTA > 250. The distributions are very different. It seems reasonable to conclude from these results that the linear part of the log frequency distribution in Figure 7 is due to real earthquakes and probably reflects Gutenberg-Richter's frequency-magnitude distribution; the nonlinear portion of the distribution probably is not due to real earthquakes but are false alarms occurring when the 0.9 to 1.4 hz noise component becomes suddenly larger in some specific beam. It appears that the frequency of false alarms is inversely proportional to the fourth power of MSTA. In Figure 9 the cumulative false alarm percent of the total number of detections is plotted versus detections with MSTA above a variable threshold.

So far the MSTA distributions have been determined for genuine signals and noise. The next parameter MSTA2 would depend on MSTA1 if both detections are related to the same earthquake. Assuming that MSTA is logarithmically related

to signal strength then ΔMSTA=MSTA1-MSTA2 would be related to the attenuation factor. From earlier studies it has been observed that the attenuation factor is very variable even for the same phase coming from the same region. The only consistent observation is that later phases are rarely stronger than the first arrival except in the shadow zone. The distribution of ΔMSTA was estimated and approximated by two exponentials.

At this point there is sufficient information available to estimate Λ_i using the transformed set of parameters and equation (2). Table 2 lists the actual distributions used to approximate the probability density functions of S1, S2, S3, MSTA1 and ΔMSTA. Time was saved by computing log Λ_i since this is just a monotonic increasing transformation.

In summary, the phase identifier works as follows. All detections that have occurred in the last half hour and that have triggered a beam from which P or PKP phases were observed in the training set are stored in a memory buffer. Any new detection is then tested with each of the earlier detections in the buffer. For a specific later phase to be considered ΔT of the detection pair must fall in a specific time window of that interpretation. Next the phase identifier would attempt to transform to new variables S1, S2, ... If the later phase was never observed in NBM2, then that interpretation would be rejected at this point.

$\text{Log}_e \Lambda$ would be computed by equation (3)

$$\log_e \Lambda = \log_e \Lambda_i - \log_e \Lambda_f \tag{3}$$

where Λ_f is computed assuming the interpretation is false. If $\log_e \Lambda$ is negative it means that it is more probable that the interpretation was incorrect and hence it is immediately rejected. On the other hand if it is positive it is now ascertained that the new detection is a later phase. The final interpretation for the new phase is the one which maximizes $\log_e \Lambda$. If all possible interpretations were rejected then the phase identifier concludes the new detection is not any of the later phases it was designed to identify. Hence the new detection is either a P, PKP, or noise. If no first arrivals were observed in NBM2 then the detection is taken to be noise, otherwise it is tentatively assumed to be a first arrival and stored in the buffer. Unless the MSTA happened to be large, one could never know for sure whether the new arrival was noise or a first arrival. The only exception is when a later arrival is discovered to correspond to the first arrival. As will be described in the next section this implementation of phase identifier appears to work very well.

<u>Performance</u> - The phase identifier described in the last section was run through 67 days of actual data. On the average it picked out 9 later phases a day. Of these 9

later phases only three of them could be confirmed by using outside bulletins. About half of these phases are so small that an analyst would probably not notice them. There was no way by which we could confirm the remaining 6 phases since corresponding earthquakes were not found in the SAAC or CGS catalogs. However, it shall be shown that it is very probable that 3 of these unconfirmed later phases were correctly identified.

The phase identifier made few errors in misclassifying a later phase. Later phase identifications in the SAAC bulletins were compared with those put out by the phase identifier. Later phases in the SAAC bulletin that did not cause a detection were not included in this analysis. The results are shown by the confusion matrix in Table 3. The row is the SAAC identification. Along any row is the distribution of phase identifier decisions. The sum of the off diagonal terms (misclassifications) is only 10% of all the SAAC identifications. The missed phases were due generally to no first arrival in the detection log either because the signal was too weak or the LASA detection system was down. In some cases later phases were missed because the phase identifier was not expecting to see the phase in that particular beam. The later phases that were not missed but misidentified were due to the confusion resulting when two earthquakes occurred around the same time and along the same

azimuth.

It is still necessary to estimate the probability of the phase identifier picking a later phase which is actually noise or a first arrival. Since it is impossible to know from outside sources that a specific detection is noise another approach had to be taken. The identifying process is so complicated that it is very difficult to estimate the number of errors using a theoretical model. The most reliable way and also the easiest way is by simulation using a synthetic detection log.

Using a random number generator a new detection log was generated with all the statistical properties of the original detection log except one. There were no later phases in the synthetic detection log, i.e. all the detections were completely independent of each other. The method used in synthesizing this log is described in Appendix A. When this log was run through the phase identifier less than two later phases were picked out a day on the average. Thus it could be concluded that 80% of the later phases found by the phase identifier are true later phases.

At the present the automated phase identification and event verification is being extended to two arrays (LASA and NORSAR) operating simultaneously. These provide better geographical coverage, greater reliability and eventually better automatic locations of events.

Publications

Chinnery, M.A. and M.N. Toksöz, P-wave velocities in the mantle below 700 km, Bull. Seism. Soc. Am., 57, 199-226, 1967.

Lacoss, R.T., E.J. Kelly and M.N. Toksöz, Estimation of seismic noise structure using arrays, Geophysics, 34, 21-38, 1969.

Shlien, S. and M.N. Toksöz, A clustering model for earthquake occurrences, Bull. Seism. Soc. Am., 60, 1765-1787, 1970.

Shlien, S., and M.N. Toksöz, Frequency-magnitude statistic of earthquake occurrences, Earthquake Notes, 41, 5-18, 1970.

Toksöz, M.N., Crustal effects on long-period chirp filters, Proceedings of the Woods Hole Conference on Seismic Discrimination, Vol. I, 20-23 July 1970.

Toksöz, M.N., M.A. Chinnery and D.L. Anderson, Inhomogeneities in the earth's mantle, Geophys. J. R. astr. Soc., 13, 31-59, 1967.

Toksöz, M.N. and R.T. Lacoss, Mode structure and source of microseisms, Science, 159, 872-873, 1968.

Toksöz, M.N. and R.A. Wiggins, Seismic arrays and the structure of the earth's interior, NEREM Record, 170-171, 1969.

Wolfe, J.C., dt/dΔ measurement of short period P waves in the upper mantle for the Western North America using LASA, S.M. thesis, M.I.T., 1970.

TABLE 1.

Covariance Matrix

H_o

	S1	S2	S3
S1	156	34	18
S2	34	88	13
S3	18	13	236

H_1

2000	103	210
103	499	-57
210	-57	10650

TABLE 2.

N(a,b) Normal Distribution
Mean a, Variance b

	True	False
S1	N(0,111)	N(10.5,1670)
S2	N(0,72)	N(1.4,728)
S3	N(0,202)	1/360
MSTA1	$\frac{1}{420} e^{-MSTA1/420}$	$375{,}000/(MSTA1)^4$
ΔMSTA	$\frac{.88}{580} e^{-\Delta MSTA/580}$ $\Delta MSTA > 0$	$\frac{375{,}000}{(MSTA1 + \Delta MSTA)^4}$
	$\frac{.12}{88} e^{\Delta MSTA/80}$ $\Delta MSTA < 0$	

TABLE 3.

Confusion Matrix

		Phase Identifier						
		PCP	SCP	PP	SKP	PKKP	P'P'	MISSED
S.A.A.C.	PCP	39	1	0	0	0	0	3
	SCP	0	11	1	1	1	0	0
	PP	0	0	16	0	0	0	4
	SKP	0	0	0	11	0	0	2
	PKKP	0	0	0	0	38	0	0
	P'P'	0	0	0	0	0	6	0

FIGURE CAPTIONS

5-1. Percent detection of CGS events magnitude m_b and greater. Right events between 31 and 90 degrees from LASA; left events between 90 and 150 degrees from LASA.

5-2. Time interval between later arrival and first arrival in seconds versus CGS distance from LASA in degrees for PcP, ScP, PP-P, PP-PKP, SKP phases.

5-3. Time interval between later arrival and first arrival in seconds versus CGS distance from LASA in degrees for P'P', PKKP-P, PKKP-PKP phases.

5-4. Epicenter location with respect to LASA for phases arriving in specified beams.

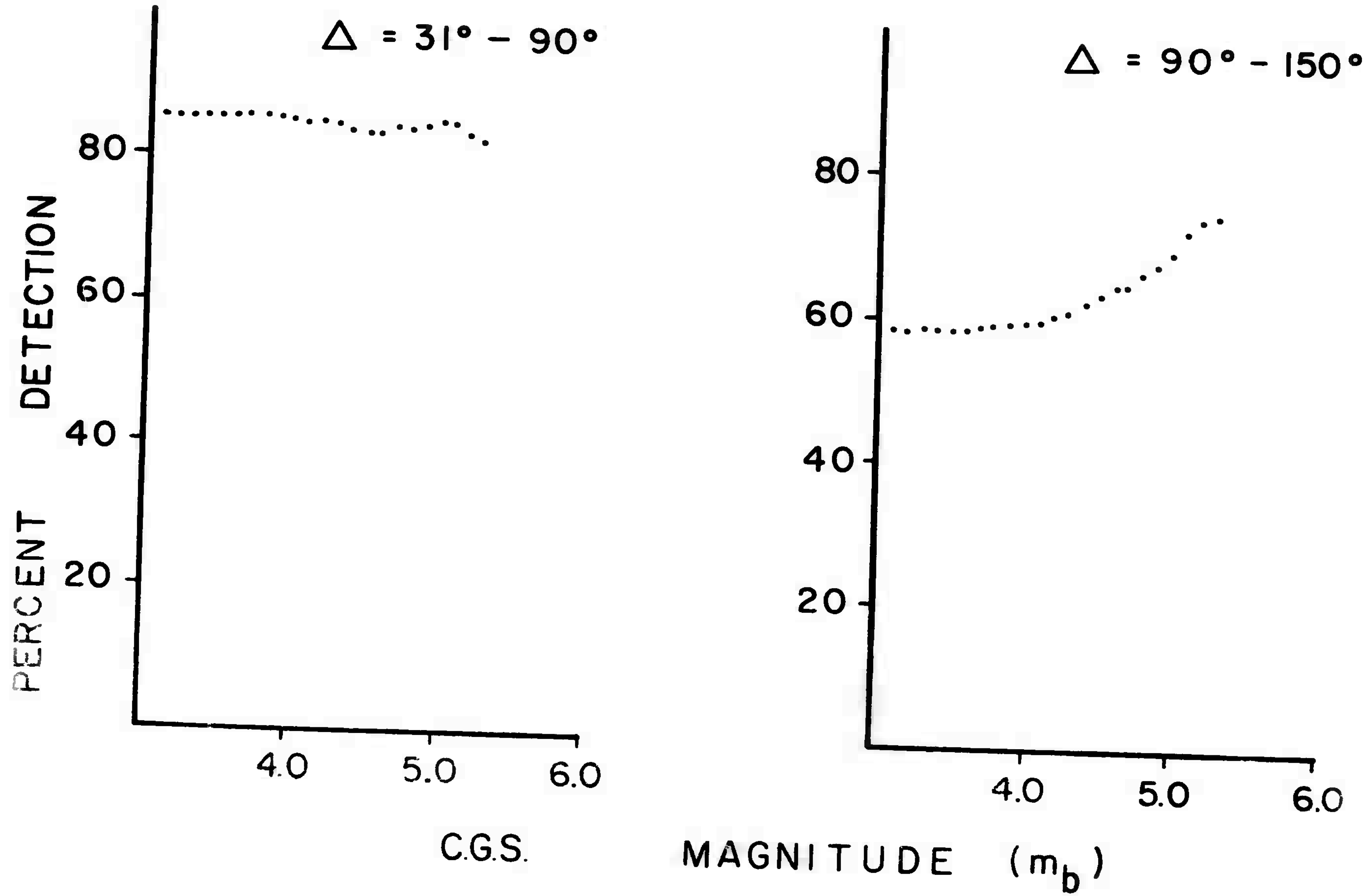
5-5. Frequency distribution of S1 in degrees. Top are phases picked out by the phase identifier (657 total samples); bottom phases from synthetic detection log (733 samples).

5-6. MSTA of detections of P phase in quantum units versus CGS magnitude. Earthquakes were shallow and confined to 30 to 60 degrees from LASA.

5-7. Cumulative number of detections versus MSTA in quantum units for 141 days from 13 May to 1 October 1969. 19253 detections with MSTA > 50.

5-8. Cumulative number of detections versus MSTA in quantum units. Upper distribution determined from 36 beams containing 61% of detections with MSTA > 250 quantum units. Lower distribution from 36 beams containing less than 0.65 of detections with MSTA > 250 over the same time interval.

5-9. Estimated false alarm rate.



Δ = 31° – 90°
Δ = 90° – 150°
PERCENT DETECTION
80
60
40
20
4.0
5.0
6.0
C.G.S.
MAGNITUDE (m_b)

Fig. 5-1

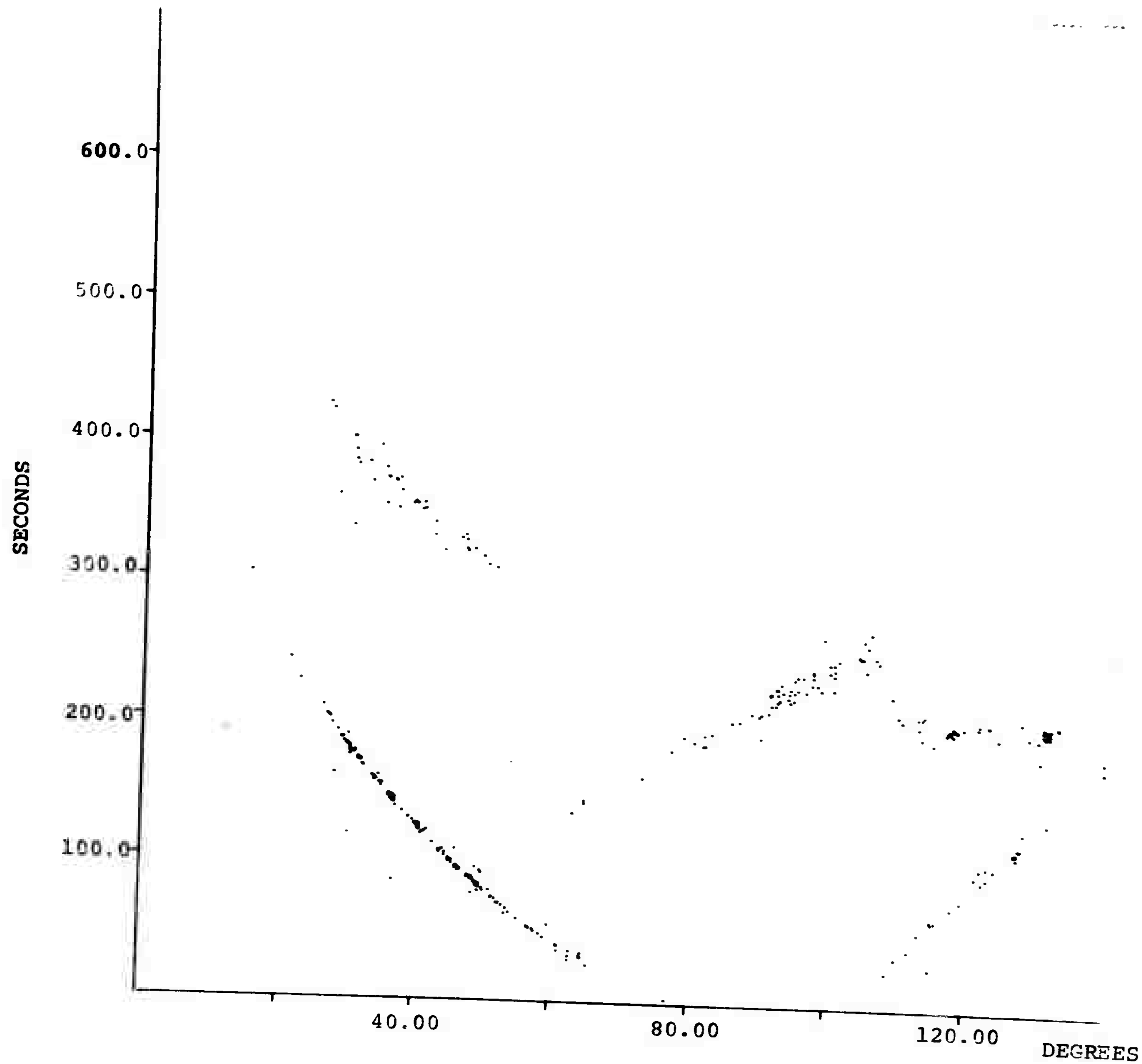

SECONDS
100.0
200.0
300.0
400.0
500.0
600.0
40.00
80.00
120.00
DEGREES

Fig. 5-2

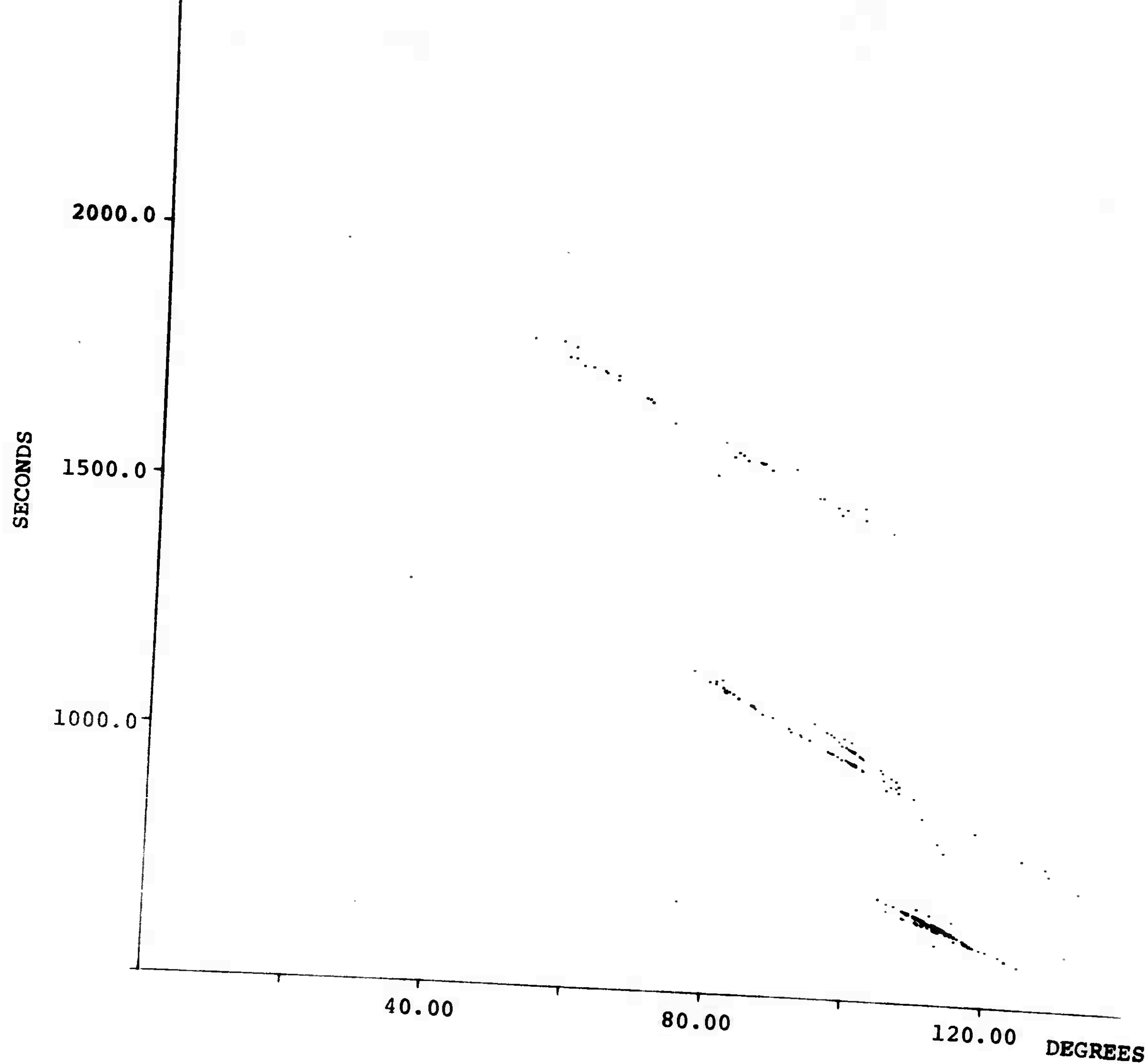


Fig. 5-3

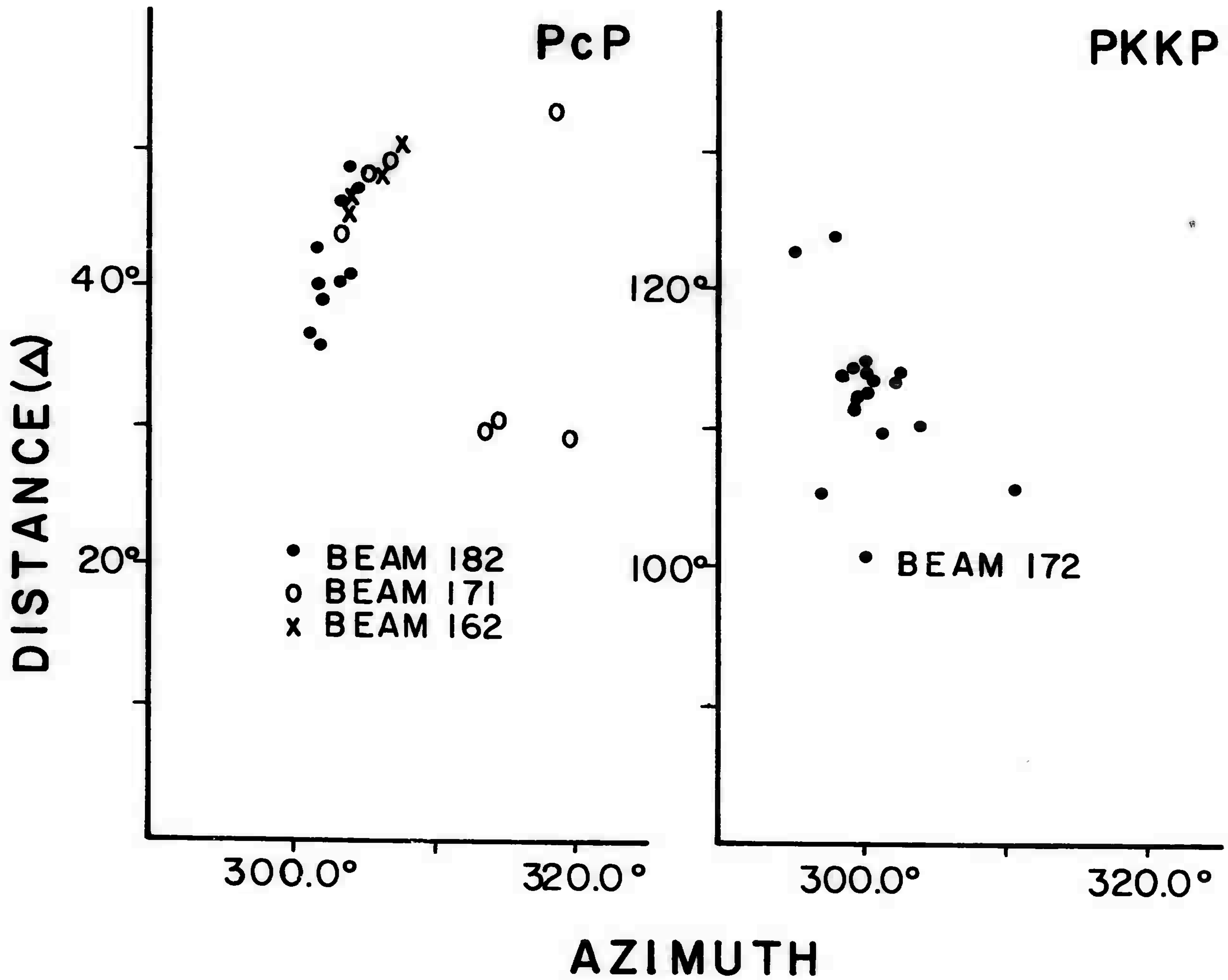

PcP
PKKP
DISTANCE (Δ)
40°
20°
120°
100°
• BEAM 182
o BEAM 171
x BEAM 162
• BEAM 172
300.0°
320.0°
300.0°
320.0°
AZIMUTH

Fig. 5-4

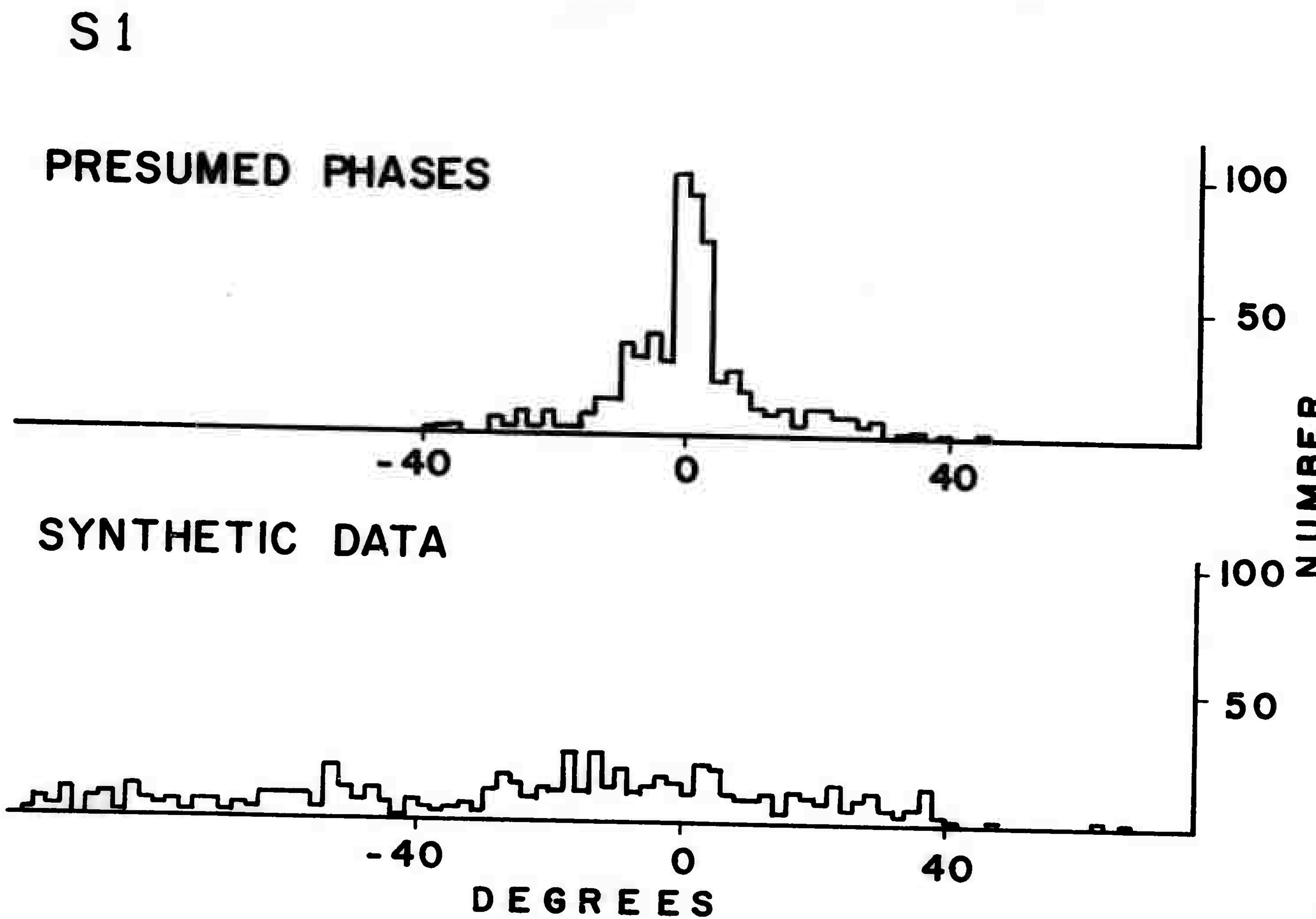

S 1
PRESUMED PHASES
SYNTHETIC DATA
100
50
-40
0
40
NUMBER
DEGREES

Fig. 5-5

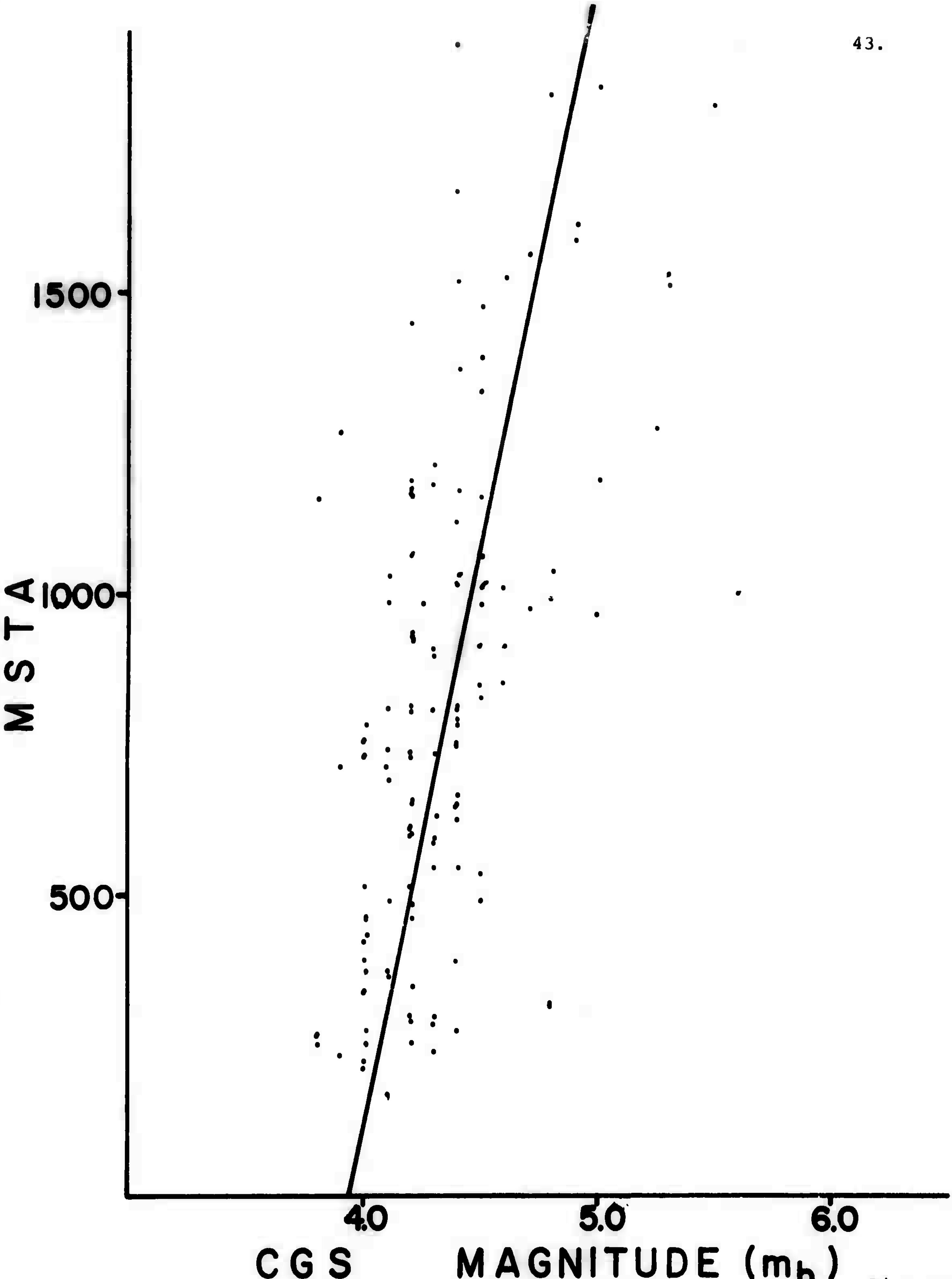
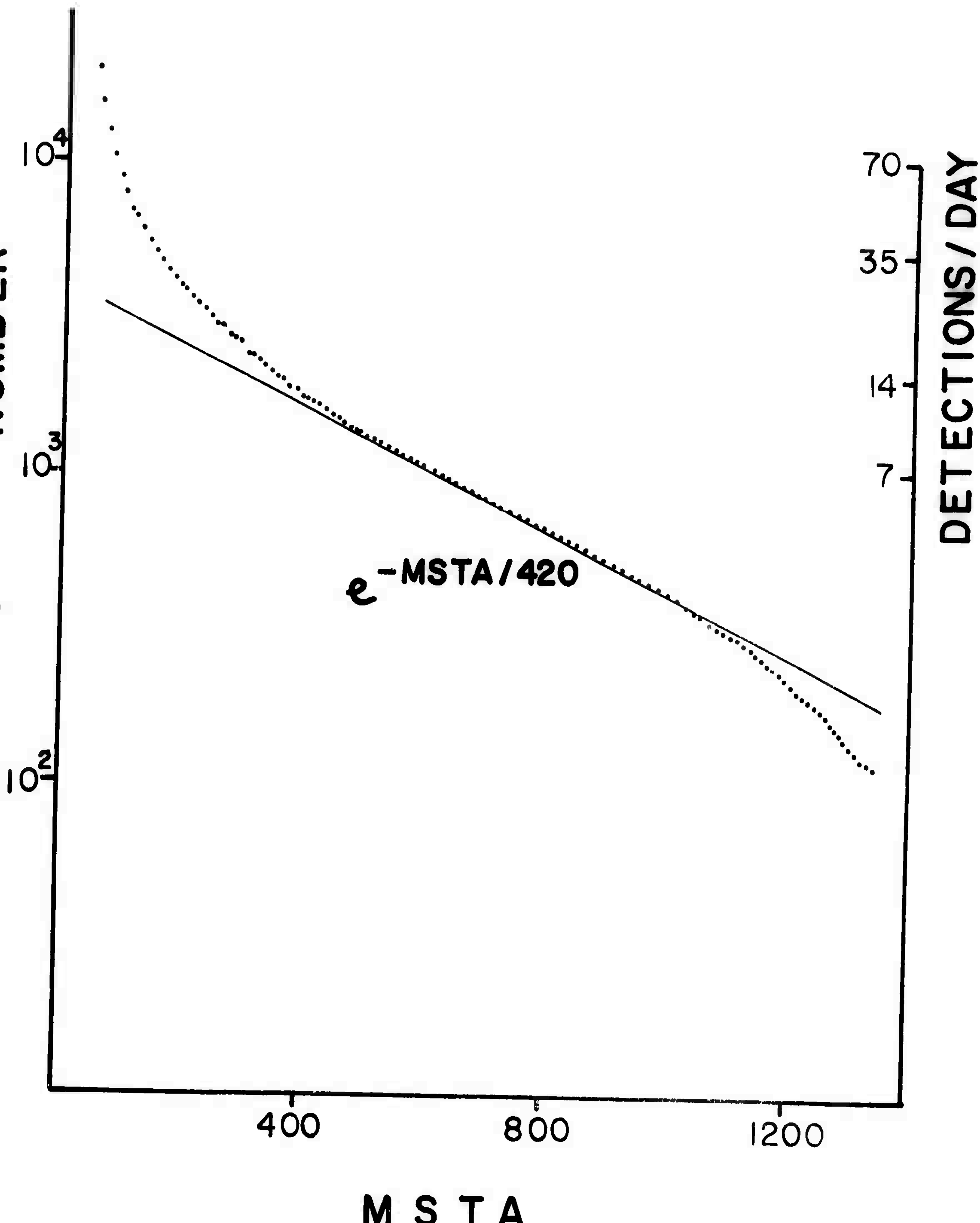


Fig. 5-(



10^4
10^3
10^2
70
35
14
7
DETECTIONS / DAY
$e^{-MSTA/420}$
400
800
1200
MSTA

Fig. 5-7

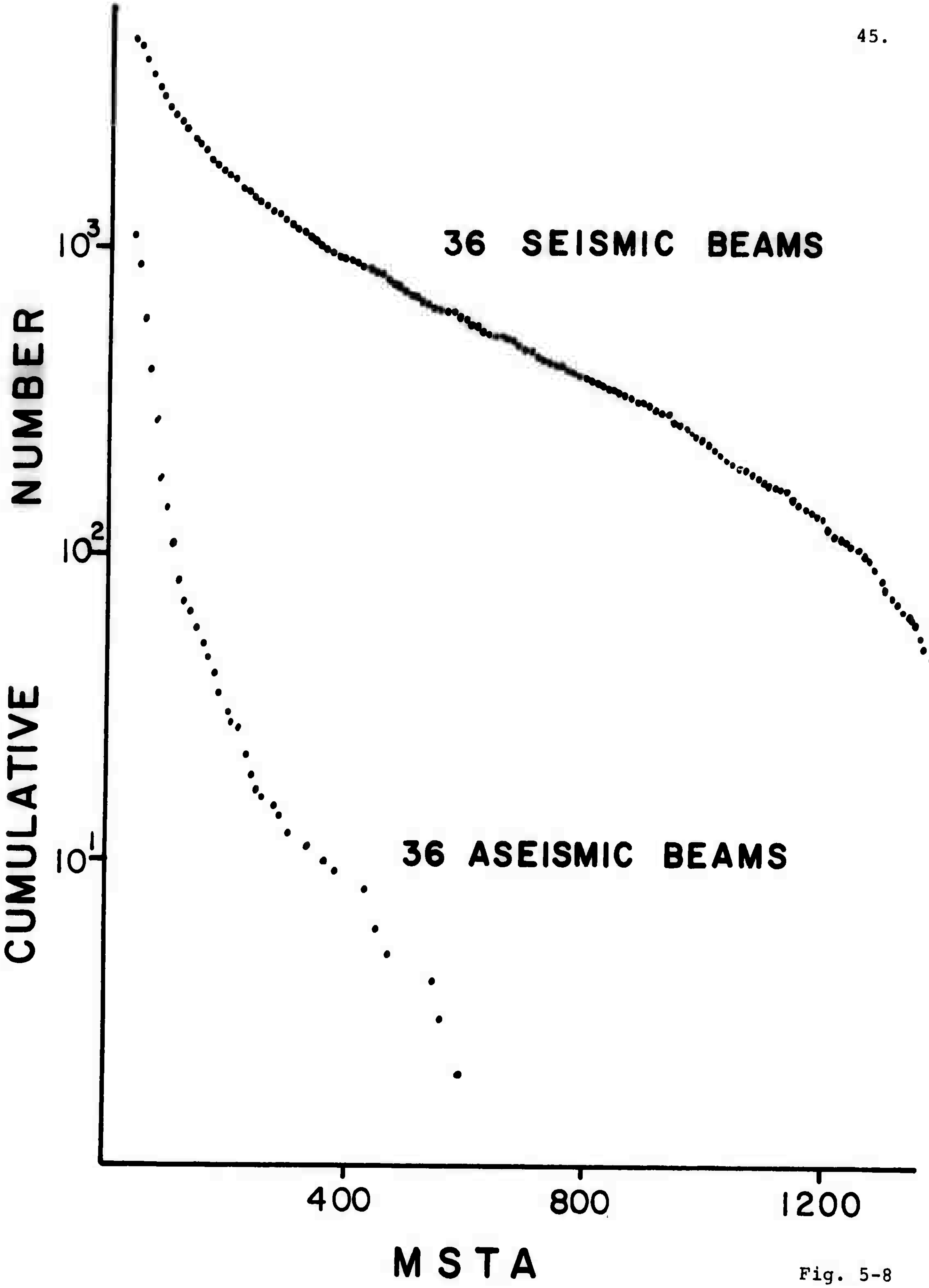

36 SEISMIC BEAMS
36 ASEISMIC BEAMS
CUMULATIVE NUMBER
10^3
10^2
10^1
400
800
1200
MSTA

Fig. 5-8

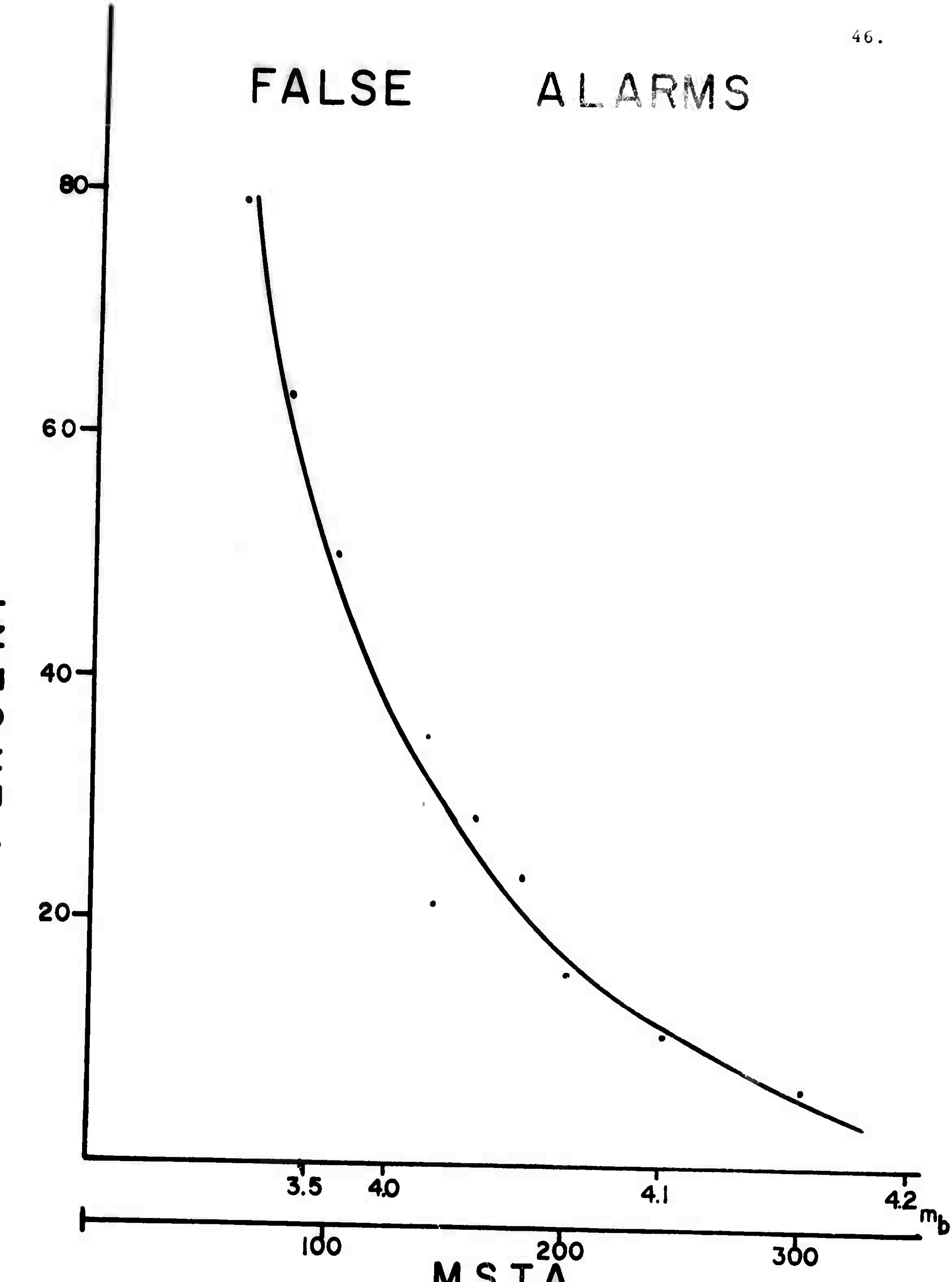


Fig. 5-9

Appendix A

In this section we describe how the synthetic detection log was generated. The synthetic detection log had to have the following properties. (1) Detections occur at completely random times, (2) About 140 detections a day have a signal strength distributed as $MSTA^{-4}$ and occur uniformly in all 300 beams, (3) About 25 detections a day are confined to seismic beams and have an exponential MSTA distribution.

The random number generator produced uncorrelated numbers uniformly distributed between plus and minus $2^{17} = 131072$ which were transformed to the interval 0 to 1. To satisfy the first condition it was assumed that detection occurrence times could be modeled by the Poisson process. Then the time intervals between adjacent detections are independently and exponentially distributed. The exponentially distributed time intervals t in seconds were generated by the following equation using the uniformly distributed random numbers x, between 0 and 1

$$t = -\frac{1}{k} \log_e(1-x) \qquad (1A)$$

k is the number of detections in a day divided by the number of seconds in a day (86,400).

For the second and third conditions another random

number was picked. If the number was below a certain threshold then the detection was taken to be a seismic signal, otherwise it was noise. The threshold was chosen so that there were about 25 seismic signals a day. If the seismic signal was noise then

$$\Pr(\mathrm{MSTA}|\mathrm{noise}) = A/\mathrm{MSTA}^4 \qquad \mathrm{MSTA} > 50 \tag{2A}$$

where the normalization constant $A = 3 \cdot 50^4 = 375000$. Another x was computed and MSTA was determined by

$$\mathrm{MSTA} = (A/3x)^{1/3} \tag{3A}$$

The beam number was determined so that it was uniformly distributed between 1 and 300.

If the signal was seismic then MSTA probability density function is given by

$$\Pr(\mathrm{MSTA}|\mathrm{seismic}) = (1/420) \cdot \exp(-\mathrm{MSTA}/420) \tag{4A}$$

Hence MSTA was determined by

$$\mathrm{MSTA} = -420\log_e(1-x) \tag{5A}$$

The beam number was determined so that it followed the seismic distribution in Table 1.

It was found that the number of later phases picked out by the phase identifier was fairly insensitive to k but depended on the seismic beam distribution.

VI. INSTRUMENTATION

Long-period tiltmeters were installed at two stations - Harvard, Mass. and Eilat, Israel - under this contract. These tiltmeters are redesigned versions of the Benioff mercury pendulum. The mechanical stability is improved by manufacturing tanks of glass. Bridge electronics operates on a modified concept with improved stability and S/N ratio. An inert, stable transformer oil-fills the gap between the mercury pool surfaces and capacitor plates.

The base length of each component of the tiltmeter at Harvard is 19 ft, and the natural period is 70 sec. Linear tilt range is 2×10^{-4} radians. Resolution is 1×10^{-9} radians. At the Israel installation the base length is increased, resulting in greater sensitivity and longer natural period (T_o = 102 sec). The resolution is 1×10^{-11} radians. At both installations the instruments record in three pass-bands: (1) Surface wave band, (2) Eigen-frequency band, and (3) D.C. output.

Because of greater sensitivity, a seismically quieter site (low level of microseisms) and better temperature stability at the vault, the performance of the tiltmeter installed at Eilat is far superior to that at Harvard station.

As of January 1971, about three months of records were available to assess the performance of the tiltmeters installed in Eilat. This is not yet a sufficient suite of

records to achieve the research goals referred to earlier, but it does provide preliminary indications on the performance of the instruments.

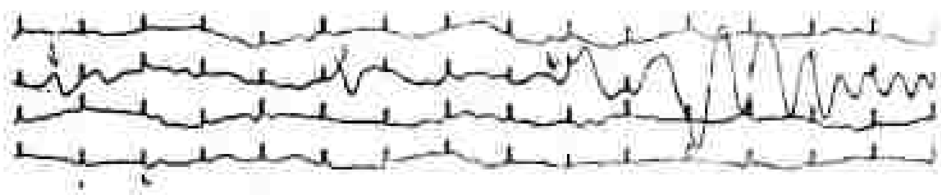
An examination of the records indicates that in the surface-wave band the instruments are operating near the maximum sensitivity permitted by the ground noise characteristic of the winter months. Widening the gap of the transducer capacitance will lead to some improvement in performance in the winter months, and to a marked improvement in the summer and on quiet winter days. The surface-wave channel should be markedly improved by this modification.

The records differ from conventional seismographs in showing mantle surface waves from relatively small events. We have not been able to identify most of these events because their magnitudes are too small to be reported through usual channels. In further studies, we will track these tremors down in order to study the efficiency of long wave excitation in the magnitude range less than 5.

Some sample records are shown in Fig. 6-1. In examining new records, the reader should remember that the response at 20 seconds is down by a factor of about 5 compared to the response at 70 seconds, so that performance of the surface-wave channel should be judged by the detection of waves with periods in the range 50-100 seconds. Other seismographs are better suited for recording surface waves

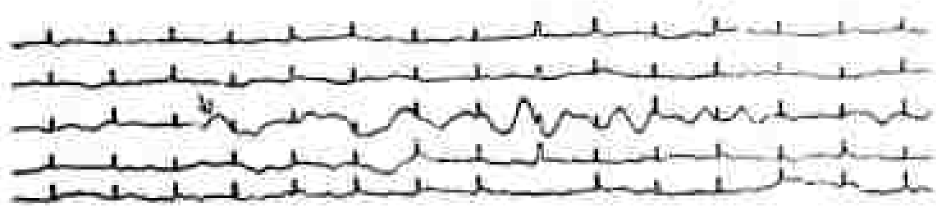
with periods under 30 seconds, and the eigenmode channel will be more useful for waves with periods exceeding 200 seconds.

The records in Fig. 6-1 are typical of a large number of events detected in the three-month operating period showing crustal-mantle surface waves containing energy near 100 seconds from earthquakes with magnitudes near 5. The background noise shown in Fig. 6-1 is typical, indicating that the gain could have been increased by two or more times about 50% of the time.



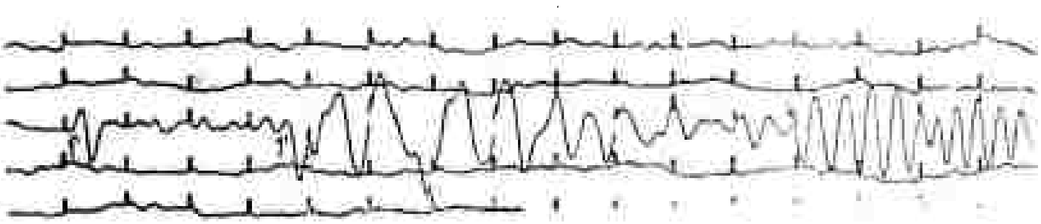
(1)

Record 1. Prince Edward Island, M = 5.1, $\Delta = 75^{\circ}$. Showing excitation 80-second surface waves on E-W component.



(2)

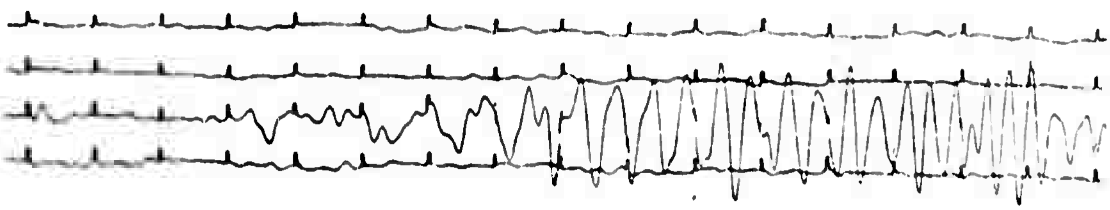
Record 2. California, M = 5.4, $\Delta = 110^{\circ}$. Showing excitation of 100-second surface waves on N-S component.



(3)

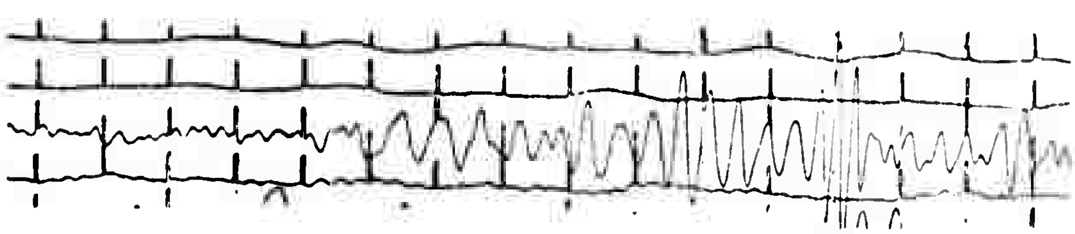
Record 3. Jan Mayen Island, M = 5.1, $\Delta = 47^{\circ}$. Showing excitation of 80-second surface waves on E-W component.

Fig. 6-1 Surface wave records



(4)

Record 4, 5. Large Soviet underground explosion 14 October 1970 on E-W (4) and N-S (5) components, maximum period about 60 seconds.



(5)

Fig. 6-1 Surface wave records (continued).

LIST OF PUBLICATIONS

Aki, K., Scaling law of seismic spectrum, J. Geophys. Res., 72, 1217-1231, 1967.

Aki, K., Seismological evidences for the existence of soft thin layers in the upper mantle under Japan, J. Geophys. Res., 73, 585-594, 1968a.

Aki, K., Seismic displacements near a fault, J. Geophys. Res., 73, 5359-5376, 1968b.

Aki, K., Analysis of the seismic coda of local earthquakes as scattered waves, J. Geophys. Res., 74, 615-631, 1969.

Aki, K. and K. Larner, Surface motion of a layered medium having an irregular interface due to incident plane SH waves, J. Geophys. Res., 75, 933-954, 1970.

Aki, K. and Y.B. Tsai, The mechanism of Love wave excitation by explosive sources, (submitted to J. Geophys. Res., 1971).

Arkani-Hamed, J., Lateral variations of density in the mantle, Geophys. J. R. Astr. Soc., 20, 431-455, 1970.

Arkani-Hamad, J., and M.N. Toksöz, Analysis and correlation of geophysical data, Supplemento al Nuovo Cimento, Ser. I, 6, 22-66, 1968.

Boore, D., Effect of higher mode contamination on measured Love wave phase velocities, J. Geophys. Res., 74, 6612- , 1969.

Boore, D., Love waves in nonuniform wave guides: Finite difference calculations, J. Geophys. Res., 75, 1512- , 1970.

Boore, D., K. Larner and K. Aki, Comparison of two independent methods for the solution of wave-scattering problems: response of a sedimentary basin to vertically incident SH waves, J. Geophys. Res., 76, 558-569, 1971.

Boore, D.M. and M.N. Toksöz, Rayleigh wave particle motion and crustal structure, Bull. Seism. Soc. Am., 59, 331-346, 1969.

Chinnery, M.A. and M.N. Toksöz, P-wave velocities in the mantle below 700 km, Bull. Seism. Soc. Am., 57, 199-226, 1967.

Fairborn, J.W., Shear wave velocities in the lower mantle, Bull. Seism. Soc. Am., 59, 1983-1999, 1969.

Frasier, C., Discrete time solution of plane P-SV waves in a plane layered medium, Geophysics, 35, 197-219, 1970.

Gangi, A.F. and J.W. Fairborn, Accurate Determination of seismic array steering delays by an adaptive computer program, Supplemento al Nuovo Cimento, Ser. I, 6, 1968.

Harkrider, D. and F. Press, The Krakatoa air-sea waves: an example of pulse propagation in coupled systems, Geophys. J., 13, 219-222, 1967.

Husebye, E. and R. Madariaga, Origins of precursors to core waves, Bull. Seism. Soc. Am., 60, 939-952, 1970.

Kanamori, H., Synthesis of long-period surface waves and its application to earthquake source studies - Kurile Islands Earthquake of October 13, 1963, J. Geophys. Res., 75, 5011-5027, 1970.

Kanamori, H., The Alaska earthquake of 1964: Radiation of long-period surface waves and source mechanism, J. Geophys. Res., 75, 5029-5040, 1970.

Kanamori, H. and F. Press, How thick is the lithosphere? Nature, 226, 330-331, 1970.

Lacoss, R.T., E.J. Kelly and M.N. Toksöz, Estimation of seismic noise structure using arrays, Geophysics, 34, 21-38, 1969.

Madariaga, R., Study of free oscillations of the laterally heterogeneous Earth models, abstract, program for Cordilleran Section, 67th Annual Meeting at Riverside, Calif., Geol. Soc. Am., 1971.

Minear, J.W. and M.N. Toksöz, Thermal regime of a downgoing slab, Tectonophysics, 10, 367-390, 1970.

Minear, J.W. and M.N. Toksöz, Thermal regime of a downgoing slab and new global tectonics, J. Geophys. Res., 75, 1397-1419, 1970.

Press, F., Free oscillations, aftershocks, and Q, in The Earth Beneath the Continents, ed. by J.S. Steinhart and T.J. Smith, AGU Monograph #10, 1966, 498-501.

Press, F. and D. Harkrider, Air-sea waves from the explosion of Krakatoa, Science, 154, 1325-1327, 1966.

Press, F., Spectra of free oscillations from an aftershock sequence, Geophys. J. Roy. Astron. Soc., 13, 219-222, 1967.

Press, F., Earth models obtained by Monte Carlo inversion, J. Geophys. Res., 73, 5223-5234, 1968.

Press, F., The sub-oceanic mantle, Science, 165, 174-176, 1969.

Press, F., Earth models consistent with geophysical data, Phys. Earth Planetary Interiors, 3, 3-22, 1970.

Shlien, S. and M.N. Toksöz, A clustering model for earthquake occurrences, Bull. Seism. Soc. Am., 60, 1765-1787, 1970.

Shlien, S., and M.N. Toksöz, Frequency-magnitude statistic of earthquake occurrences, Earthquake Notes, 41, 5-18, 1970.

Solomon, S.C. and M.N. Toksöz, Lateral variation of attenuation of P and S waves beneath the United States, Bull. Seism. Soc. Am., 60, 819-838, 1970.

Solomon, S.C., R.W. Ward and M.N. Toksöz, Earthquake and explosion magnitudes: The effect of lateral variation of seismic attenuation, Proceedings of the Woods Hole Conference on Seismic Discrimination, Vol. I, 20-23 July, 1970.

Solomon, S.C., Seismic-waveaattenuation and partial melting in the upper mantle of North America, J. Geophys. Res., in press, 1971b.

Toksöz, M.N., Radiation of seismic waves from underground explosions, Proceedings of the Vesiac conference on the current status and future prognosis of shallow seismic events, VESIAC Report, Willow Run Laboratories, The University of Michigan, 65-84, 1967.

Toksöz, M.N., and J. Arkani-Hamed, Seismic delay times: correlation with other data, Science, 158, 783-785, 1967.

Toksöz, M.N., M.A. Chinnery and D.L. Anderson, Inhomogeneities in the earth's mantle, Geophys. J. R. astr. Soc., 13, 31-59, 1967.

Toksöz, M.N., and R.T. Lacoss, Mode structure and source of microseisms, Science, 159, 872-873, 1968.

Toksöz, M.N., J. Arkani-Hamed and C.A. Knight, Geophysical data and long wave heterogeneities of the earth's mantle, J. Geophys. Res., 74, 3751-3770, 1969.

Toksöz, M.N. and R.A. Wiggins, Seismic arrays and the structure of the Earth's interior, NEREM Record, 170-171, 1969.

Toksöz, M.N., Crustal effects on long-period chirp filters, Proceedings of the Woods Hole Conference on Seismic Discrimination, Vol. I, 20-23 July 1970.

Toksöz, M.N. and H.H. Kehrer, Underground nuclear explosions: tectonic utility and dangers, Science, 173, 230-233, 16 July 1971.

Toksöz, M.N., J. W. Minear and B.R. Julian, Temperature field and geophysical effects of a downgoing slab, J. Geophys. Res., 76, 1113-1138, 1971.

Tsai, Y.B. and K. Aki, Simultaneous determination of the seismic moment and attenuation of seismic surface waves, Bull. Seism. Soc. Am., 59, 275-287, 1969.

Tsai, Y.B. and K. Aki, Precise focal depth determination from amplitude spectra of surface waves, J. Geophys. Res., 75, 5729-5743, 1970.

Tsai, Y.B. and K. Aki, Amplitude spectra of surface waves from small earthquakes and underground nuclear explosions, J. Geophys. Res., 76, 3940-3952, 1971.

Ward, R.W. and M.N. Toksöz, Causes of regional variation of magnitude, Bull. Seism. Soc. Am., 61, 649-670, 1971.

Wiggins, R. and F. Press, Search for seismic signals at pulsar frequencies, J. Geophys. Res., 74, 5351-5352, 1969.

LIST OF THESES

Masters

Harita, Y., An investigation of the upper mantle P-velocity models using synthetic seismograms, S.M. thesis, M.I.T., 1969.

Horowicz, L., Two methods for separating higher modes of seismic surface waves, S.M. thesis, M.I.T., 1966.

Kuster, G.T., Sound velocity in fluid-solid suspension, S.M. thesis, M.I.T., 1970.

Wolfe, J.C., dt/dΔ measurement of short period P waves in the upper mantle for the Western North America using LASA, S.M. thesis, M.I.T., 1970.

Ph.D.

Arkani-Hamed, J., Lateral variations of density in the earth's mantle, Ph.D. thesis, M.I.T., 1969.

Boore, D., Finite Difference solutions to the equation of elastic wave propagation, with application to Love waves over dipping interfaces, Ph.D. thesis, M.I.T., 1970.

Fairborn, J.W., Mantle P and S wave velocity distributions from dt/dΔ measurements, Ph.D. thesis, M.I.T., 1968.

Frasier, C., Discrete time solution of plane P-SV waves in a plane layered medium, Ph.D. thesis, M.I.T., 1969.

Larner, K., Near-receiver scattering of teleseismic body waves in layered crust-mantle models having irregular interfaces, Ph.D. thesis, M.I.T., 1969.

Madariaga, R.I., Free oscillations of the laterally heterogeneous earth, Ph.D. thesis, M.I.T., 1971.

Solomon, S.C., Seismic wave attenuation and the state of the upper mantle, Ph.D. thesis, M.I.T., 1971a.

Tsai, Y.B., Determination of focal depths of earthquakes in the mid-oceanic ridges from amplitude spectra of surface waves, Ph.D. thesis, M.I.T., 1970.